Tracer evolution in winds generated by a global spectral mechanistic model

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Abstract.
The lower boundary of a spectral mechanistic model is prescribed with 100 hPa geopotentials, and its performance during a November 1989 through March 1990 integration is compared with National Meteorological Center observations. Although the stratosphere temperatures quickly become biased near the pole in both hemispheres, the model develops a residual mean circulation which shows significant descent over the winter pole and ascent in the tropics and over the summer pole at pressures less than 10 hPa. The daily correspondence of observed to modeled features in the upper stratosphere and mesosphere degrades after one month. However, the long-term variability qualitatively follows the observations. The results of off-line transport experiments are also described. A passive tracer is instantaneously injected into the flow over the poles and evolves in a manner which is consistent with the residual mean circulation. It demonstrates a significant cross-equatorial flux in the mesosphere near solstice, and air which originates in the stratospheric polar mesosphere can be found descending deep into the northern polar stratosphere at the end of the integration. Nitrous oxide is also transported, and its ability to act as a dynamical tracer is evaluated by comparison to the evolution of the passive tracer.

Introduction

Because of its long photochemical lifetime in much of the stratosphere and its lack of sources there, nitrous oxide $(N_2O)$ has long been recognized as a potentially powerful indicator of the transport properties of the middle atmosphere. Within the darkness of polar night, even destruction is negligible, so $N_2O$ is expected to be a nearly perfect tracer of motion within the winter polar vortices. While participating in both the Airborne Antarctic Ozone Experiment (AAOE) and the Airborne Arctic Stratospheric Expedition (AASE), Loewenstein et al. [1989, 1990] examined $N_2O$ mixing ratios across the edge of each polar vortex. Assuming that the vortex wall acts as a container, with little diffusion or mixing across it [Schoeberl et al., 1992], then, at 17 km altitude, the $N_2O$ gradient at the vortex edges indicated that air within each vortex had descended 4-6 km during each respective winter. Toon et al. [1992] deduced similar rates of descent in the northern polar vortex by examining other trace constituents. The results of these efforts established the importance of long-lived tracer measurements as a tool to understand atmospheric motions.

Quantifying the magnitude of polar descent is important since it determines whether polar stratospheric clouds may form in the vortex and whether ozone can then be destroyed on the particulate surfaces by heterogeneous chemistry. Observations are also being used to investigate other related questions. How is the ozone-depleted air dispersed at the end of the Antarctic winter? Can the long-term evolution of a tracer indicate the extent of the exchange across “hydrodynamical barriers,” including not only the polar vortex edge but also the tropopause? How rapidly is air passed between the tropics and the middle latitudes? How do hemispheric asymmetries influence interhemispheric transport? What is the forcing of the mean meridional circulation from gravity wave drag? Tracer constituent information returned by the Upper Atmosphere Research Satellite (UARS) and from future Earth observation platforms and ground-based campaigns will provide some answers to these questions.

However, such observational information is spatially and temporally limited. To obtain a global perspective, and to isolate and quantify transport mechanisms, numerical modeling of trace constituents is required. Unfortunately, three-dimensional simulations of the observed distribution of stratospheric $N_2O$ within and near the polar vortices have previously yielded disappointing results. Parrish et al. [1988] reported that...
there were no transport or chemical modeling results which could explain the depressed N₂O mixing ratios which they observed at McMurdo, Antarctica (78°S, 167°E). To date, perhaps the most comprehensive general circulation model (GCM) based three-dimensional transport experiments have been described by Mahlman et al. [1986] and Plumb and Mahlman [1987]. After lengthy simulations, they noted that the N₂O mixing ratio contours, in the zonal average, were sloping toward the poles by one third less than is observed. This deficiency was explained, in general, by the lack of wave-mean flow interaction and, more specifically, by the inability of the GCM to produce sudden stratospheric warmings. In effect, the model's winter poles become dominated by intense, cold polar vortices in which descent is underestimated. Recent studies by Strahan and Mahlman [1994] have shown that improvement can be expected if the depth of the model is extended and its horizontal resolution is increased. Their simulations included two northern winters in which many features of the descent of N₂O were more reasonably represented. However, their southern winter vortex was still persistently cold.

In other recent transport studies, winds from data assimilation have been successful in representing the synoptic and planetary scale variations in the upper troposphere and lower stratosphere [Rood et al., 1989, 1992, and references therein]. The isolation of the polar vortex has been well simulated. Additionally, formation of “cold” poles has been prevented by forcing the assimilation cycle to be strongly influenced by satellite or radiosonde retrievals. But inconsistencies in the diabatic balance which arise from the data insertion and the subsequent adjustment processes have led to poor performance of the transport in the middle and upper stratosphere when integrations extend beyond a few weeks [Weaver et al., 1993].

In order to address some of these problems, several investigators have adopted a “mechanistic” approach, which assumes that planetary waves in the stratosphere originate in the troposphere and propagate upward. The wave information is transmitted into the model through the bottom boundary which, in most applications, is isobaric and often lies near the tropopause. The geopotential of this surface is specified. For experiments in which theoretical considerations are important, an analytic wave spectrum, either stationary or transient, is imposed. O’Neill and Pope [1988] studied the response of their model stratosphere upon application of idealized perturbations to axially symmetric flow at 300 hPa. Kouker and Brasseur [1986] studied tracer transport in a stratospheric warming which occurred in their hemispheric semispectral model after the application of an analytic wavenumber-one perturbation. Austin and Butchart [1992] examined polar ozone photochemistry while varying the amplitude of a stationary wave at 316 hPa.

However, for experiments described here, in which we examine a specific time period, the bottom boundary is generated from archived global geopotentials. This guarantees continuous introduction of the observed planetary and synoptic scale wave activity. Rather than restricting idealized waves to grow upon a zonally symmetric state, the use of this forcing promotes interacting vortices as the waves propagate vertically. Fairlie et al. [1990] used the United Kingdom Meteorological Office 100-hPa heights to generate stratospheric warmings and small-scale structure in their gridded mechanistic model. Farrara et al. [1992] used National Meteorological Center (NMC) 100-hPa geopotentials to examine the dynamics of transient planetary waves in the early wintertime southern stratosphere.

In this paper we present results from a global spectral mechanistic model which was developed to study the dynamics of, and constituent transport in, the middle and upper stratosphere and the lower mesosphere. We focus on the results of experiments in which winds and temperatures from the mechanistic model were interfaced with a gridded chemistry and transport model (CTM). The transport integrations included a passive tracer, whose initial condition was configured to represent an instantaneous perturbation from a solar proton event (SPE), and N₂O which was given a realistic and balanced initial condition and was photochemically active. The former serves primarily as a test of the dynamics of the mechanistic model both for mixing of the perturbation and for long-term transport characteristics. The latter will eventually be compared to UARS observations and thus tests both the dynamics and the parameterization of the photochemistry.

We have already used the models described below to investigate perturbations to trace constituents following the October 1989 SPEs. Those calculations, which included the major reactions involving the oxygen, hydrogen, and nitrogen species, were reported by Jackman et al. [1993]. Interhemispheric differences in ozone depletion which were observed two months afterward could largely be attributed to differences in the transport produced by the seasonal transitions which were in progress at the time of the SPEs.

Both the mechanistic model and the CTM are described in the following section. In section 3 the initial conditions for the dynamics and for both the passive tracer and N₂O are outlined. In section 4 we assess the performance of the mechanistic model, with a brief comparison to NMC winds and temperatures, and discuss the evolution of the tracer and N₂O. Section 5 presents a summary and some concluding remarks.

The Numerical Models

Global Spectral Mechanistic Model

The global spectral mechanistic model (GSMM) is derived from the National Center for Atmospheric Research Community Climate Model, version 0 (CCM0) [Washington, 1982]. The important features of CCM0 are described by Pitcher et al. [1983]. Several modifications are introduced to accommodate or simplify integrations for the stratosphere and lower mesosphere.
The lower boundary, or reference surface, is formulated so that it lies at a constant pressure. The reference surface heights are updated at each model time step and are derived from the daily NMC global 100-hPa geopotentials. To obtain the heights between observations, the two temporally nearest geopotential fields are interpolated linearly in time. The equation for the surface pressure tendency is eliminated. The vertical velocity at the reference surface is computed by downward integration of the divergence and is, in general, not zero. Thirty constant pressure levels, evenly spaced in the natural logarithm of pressure, were chosen for this study. The uppermost level lies at 0.01 hPa, and the vertical resolution is ~2.25 km. The Legendre formulation was changed from rhomboidal truncation to triangular, and the spectrum was truncated at 351 harmonics. This corresponds, in the traditional notation, to T25 resolution. The gridded field variables are then obtained at eighty evenly spaced longitudes on each of sixty-four Gaussian latitudes.

Heating and cooling rates were calculated using the radiative transfer algorithms described by Rosenfield et al. [1987]. Included was the wide band parameterization of ozone infrared absorption at 9.6 microns which was introduced by Rosenfield [1991]. The necessary ozone and water vapor mixing ratios are not predicted in the mechanistic model. Rather, they are obtained from the monthly and zonally averaged data sets described by Rosenfield et al. [1987]. These were then interpolated onto the mechanistic model's physical space grid. Climatological water vapor was not available at pressures less than 1 hPa. The upper stratospheric and mesospheric profiles were thus obtained by linearly interpolating between the local values at 1 hPa and a constant 1.6 parts per million by volume at the model's uppermost level [Bevilacqua et al., 1987].

A crude approximation to the troposphere was required for calculation of reasonable radiative forcing in the model's lower stratosphere. At pressures greater than that of the reference surface, temperatures were provided by NMC daily global analyses. Surface pressure was calculated by hydrostatically integrating downward to the local topography. Surface albedo was obtained from the monthly climatological means used in the Goddard data assimilation model. Cloudiness was not considered.

Rayleigh friction serves as a parameterization of momentum dissipation due to breaking gravity waves. It is expressed as an exponential function of height:

$$ R = 2.3 \times 10^{-9} \exp \left( -\frac{3}{4} \ln \sigma \right), $$

where \( \sigma = p/1000 \) and \( p \) is the local pressure in hPa. \( R \) has units sec\(^{-1}\). Smith and Lyjak [1985], upon examining Limb Infrared Monitor of the Stratosphere (LIMS) data, speculated that Rayleigh damping rates in most GCMs probably underestimate momentum dissipation in the middle and upper stratosphere. Thus, poleward of 30° N and centered at 1 hPa, we have increased the friction beyond the exponential profile which is applied globally. The enhancement, which has a maximum value of \( 1.0 \times 10^{-6} \) sec\(^{-1}\), is active during the northern hemisphere’s midwinter months and is turned on gradually during November and off during March. Figure 1 illustrates the December global configuration of the friction coefficient along with a profile for 60° N latitude. Admittedly, this choice was made in an ad hoc manner and may not be an optimal representation. However, it considerably reduces the strength of the polar night jet, which became unstable in earlier experiments without the dissipation enhancement.

Vertical diffusion and all physical processes associated with water vapor were eliminated. Only dry convective adjustment was retained, and it has been restructured following the scheme invoked in CCM version 1 [Williamson et al., 1987]. With no vertical heat transport from tropospheric convection, temperatures near the tropical tropopause, which lies within the model domain, become unacceptably cold. In this region, Newtonian heating is invoked if the temperature falls under 185 K. It takes the form \( H = \alpha (185 - T) \) where \( T \) is the local temperature and, with \( \mu = \) colatitude,

$$ \alpha = \sin (3\mu - \pi) \text{ day}^{-1}. $$

The active region is defined by \( \pi/3 \leq \mu \leq 2\pi/3 \) and the elevation \( z \leq 20 \) km. Otherwise, \( \alpha = 0. \) This restricts the heating to within 30° of the equator and to the lowest two model levels.

Chemistry and Transport Model

The GSMM generates the winds and the temperatures which drive the off-line CTM. That is, there is no feedback between the evolving constituents and the radiation or the dynamics. The gridded data are obtained by transforming the time-averaged spherical harmonics of each field variable into the gridded physical domain. Afterward the fields are interpolated from the mechanistic model’s Gaussian latitudes onto the transport model’s set of evenly spaced latitudes. The latter have been chosen to provide approximately the same horizontal resolution as the Gaussian grid. No interpolations are required in the vertical direction. Time-averaged winds provide a better description of the circulation over finite temporal intervals than do instantaneous winds, as suggested by Mahlman and Mozim [1978]. The averaging interval is six hours.

The CTM is described by Allen et al. [1991]. Time-averaged winds from the GSMM are interpolated onto the CTM’s staggered horizontal grid. Linear interpolation between the two nearest six-hourly mechanistic model history records is used to obtain current winds and temperatures. Vertical winds are internally calculated by the kinematic method at each time step, using an algorithm which ensures continuity with the transport numerics.

The constituent continuity equation is solved by process splitting [Rood, 1987, Rood et al., 1992]. Each directional component of the advection is calculated prior to the chemical production and loss. Van Leer’s scheme,
which is easily time split and is monotonic and upstream biased [Allen et al., 1991], is used in the horizontal. Frather's [1986] upstream-biased method, which is nearly nondiffusive and conserves both mass and the first and second moments, is used in the vertical.

Initial conditions

Winds and Temperatures

The dynamical initial condition is obtained from two sources. At pressures greater than 0.40 hPa, NMC winds and temperatures from a selected date are balanced [Randel, 1987] and interpolated onto the mechanistic model's physical grid. At lower pressures the winds and temperatures are vertically interpolated between their local values at 0.40 hPa and the corresponding month's zonally averaged climatology [Fleming et al., 1988] at the uppermost level, 0.01 hPa. We must therefore allow the "fabricated" region to thermally adjust and the planetary waves to vertically propagate before the winds are employed in the chemistry and transport experiments. For the simulations described below, the date of initialization of the dynamics is 1200 UTC October 20, 1989. The period is characterized with weak planetary wave amplitudes throughout the stratosphere and mesosphere. The lid temperature at the north pole adjusted to within 2 K of its mean November value within eleven days. During the same period, the polar night jet rapidly intensified to 75 m sec^{-1}. Further increases averaged near 1 m sec^{-1} day^{-1} through mid-November. The transport experiments were thus begun at 0000 UTC November 1, 1989.

Passive Tracer

To simplify interpretation of our SPE simulations [Jackman et al., 1993], we initialized a passive tracer as a substitute for a perturbed constituent. The regions of the atmosphere which are most strongly influenced during SPEs lie within ~30° geomagnetic latitude of each geomagnetic pole. So in these two regions the mixing ratio of a passive tracer was initialized to 10 (arbitrary units) at each CTM model level, and elsewhere it was set to zero. Thus the transport experiment began with the passive tracer configured in cylindrical columns, one in each hemisphere. Each was instantaneously injected into the CTM at 0000 UTC November 1, 1989. They are subject to extremely rapid mixing and distortion throughout the early stages of the experiment. The positioning of the "cylinders" is illustrated in Figures 2a and 2b, and the corresponding zonal average is displayed in Figure 2c. The edges of the cylinders are
Figure 2. Initial condition for the passive tracer, identical at each model level, at 0000 UTC November 1, 1989. (a) Northern hemisphere and (b) southern hemisphere, with darkened region showing the positioning of the "cylinders." (c) The zonal average. Units are arbitrary and the contours are 0.5, 1, 2, 3, ..., 10.
jagged since the geomagnetic latitudes are not coincident with the transport model latitudes, and the initialization algorithm assigns zero or nonzero concentrations point-by-point.

Nitrous Oxide

Nitrous oxide ($N_2O$) is initialized by mapping representative profiles from the Goddard two-dimensional chemistry model onto the CTM's three-dimensional grid using the technique described by Douglass et al. [1990]. This method uses the vorticity and potential temperature from the mechanistic model to dictate the initial distribution of the $N_2O$ and to minimize the adjustment or "spin-up" period. $N_2O$ profiles from the two-dimensional model are fit to polynomial functions at several latitudes. In the northern hemisphere, the absolute vorticity from an early November 1989 ten-day time average is used to characterize the boundaries of Arctic, middle latitude, and tropical air. Within these well-defined regions, horizontal and vertical partitioning is then governed completely by the potential temperature. In the "transition regions," which lie between the boundaries of each characterizable air mass, the horizontal partitioning is based upon the ten-day average vorticity isolines as they are configured at initialization, 0000 UTC November 1, 1989. Vertical partitioning remains based on potential temperature. In the southern hemisphere, the circulation is making a transition to the summertime easterlies during November, and the characterization algorithm did not yield satisfactory results. The horizontal partitioning south of the equator is based on latitude.

Above 60 km, $N_2O$ which was initially mapped from the two-dimensional model to the global domain did not decrease with altitude as rapidly as was expected from the chemical loss rates. So an exponential decay as a function of height, based on the two-dimensional model's photochemical loss coefficients, was implemented as a correction. Near the reference surface at latitudes within 30° of each pole, the initial mixing ratios are ~20 parts per billion by volume (ppbv) higher than the climatology. The mapping algorithm partitions the polar caps to the Arctic regime, and the profiles within the region are then assigned based only upon potential temperature. The meridional slope of the potential temperature contours is shallow at initialization, and this reflected in the zonally averaged $N_2O$ contours. The initial zonally averaged $N_2O$ is illustrated in Figure 3.

$N_2O$ is destroyed by a parameterization which includes photolysis and reaction with $O(1D)$. The loss coefficients were derived from the two-dimensional model

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Figure 3. Initial condition for $N_2O$, 0000 UTC November 1, 1989. Units are ppbv, and the contour interval is nonuniform.
outlined by Douglass et al. [1989] and are updated at ten-day intervals. The average January N₂O loss rates are shown in Figure 4.

To determine constituent fluxes through the CTM’s lower boundary, a fictitious level is located beneath the reference surface. Mixing ratios on this surface are initialized identically to those on the reference surface and are kept constant. This implies time-varying localized sources or sinks located at the lower boundary, depending on the direction and magnitude of the vertical wind vector on the reference surface. The effect becomes noticeable in the evolution of the passive tracer, particularly in the CTM’s lowest two layers. Where the cylinders are initialized, mixing tends to create large vertical gradients between the reference surface and the artificial layer. Any positive vertical motion there will then initiate a flux of tracer material into the model. On the other hand, the initial N₂O configuration is dynamically balanced, and N₂O is usually horizontally stratified. Thus the effect the fictitious layer on its evolution in the lower levels is not immediately noticeable.

Discussion of Results
Winds and Temperatures

Between 100 and 0.4 hPa a general picture of the performance of the mechanistic model can be obtained by comparison of model statistics to NMC daily observations. At the mandatory levels, the latter have been balanced and interpolated onto a horizontal grid which has a resolution close to that of the Gaussian grid. Temperatures at 0.4 hPa may be suspect [Randel, 1992] but are included for completeness. Above 10 hPa the temperature analysis is based entirely on satellite retrievals, while below 10 hPa some radiosonde data is included.

We will show a few basic diagnostics in time series which cover the entire five months of integration.

The mean zonal wind component, the zonally averaged temperatures, and the amplitudes of geopotential wave-one at 60° S are shown in Figure 5. In November the model properly reproduces the decreasing wave amplitudes associated with the southern hemisphere’s winter to spring transition. The zero wind line descends to 50 hPa by late December and begins to ascend in late January. Waves propagating from below are absorbed at the zero wind line, and, as a result, there is little planetary wave activity in the southern summertime stratosphere. The model then properly simulates the transition from easterlies to westerlies in middle and late February during the summer to autumn transition. By the end of March, at this latitude, mean zonal wind speeds reach 40 m sec⁻¹ at 1 hPa in both the model and in the NMC data.

Keeping in mind the uncertainties in the NMC upper stratospheric temperatures, the modeled stratopause temperatures at 60° S on November 1, 1989, only eleven days after initialization, are several degrees warmer. The dynamical consequence is that the axis of the strongest easterlies, as a function of height, is 15°-20° too close to the equator within ~10 km of the stratopause. In late January, stratopause temperatures begin falling in both data sets, but the rate is faster in the NMC data, especially during March. In the lower stratosphere, near the lower boundary, the model is consistently cold. The deviations are largest in January.

Time series for 60° N are shown in Figure 6. Throughout the integration the circulation in the northern hemisphere is dominated by wave events, with the first occurring in late November. The model captures most of these events, but restoration toward a zonally sym-

Figure 4. January N₂O loss rates in log day⁻¹. Contour interval is nonuniform.
Figure 5. Time series of (top) mean zonal wind (m sec\(^{-1}\)), (middle) temperature (K), and (bottom) amplitude of geopotential height wave-one (geopotential km) for 60° S. Mechanistic model on the left and NMC daily observations on the right.

Metric state is not always as rapid as the observations indicate. In November, soon after initialization, the northern polar stratosphere has already become cold, and the mean zonal winds at this latitude are nearly 20 m sec\(^{-1}\) too strong. Although the temperature bias is not remarkable at 60° N, time averages for November and December indicate that it reaches -15 K at the stratopause at the pole. As the Rayleigh friction is enhanced in the northern upper stratosphere during November, the winds begin decreasing. Meanwhile, the modeled temperature tendencies follow the NMC data. Both the model and the data then show a chiefly wave number one event in late November. The model amplifies the wave faster, generating a 2 geopotential
km (gipkm) perturbation at 0.2 hPa before December 1. The NMC data indicate that the wave amplitude is greater during the first week in December.

In the atmosphere at 3 hPa, wave activity quickly declines between December 5-10, 1989, and wave-one amplitudes fall to less than 0.70 gipkm. Then a brief burst pushes the perturbation above 1 gipkm on December 14, 1989. In the model, wave amplitudes are declining on December 10, 1989, but not as fast as in the data. There is an indication that the model tries to imitate the pulse on December 14, 1989, but it is not so prominent with the already high amplitudes. The model waits until December 20, 1989, before its wave activity declines. Thus, while the atmosphere establishes winds in excess of 60 m sec$^{-1}$ at 1 hPa by the middle of December, the mechanistic model has weak mean zonal winds until late in the month.

The next significant event occurs in the middle of January, and it is captured by the model, although amplitudes above 1.5 gipkm do not appear below 1 hPa. Af-
terward the mean zonal wind recovers fairly rapidly and is comparable to or slightly greater than that shown in the NMC data. Then in late January and early February another event is seen. The model also simulates this and, in fact, shows a wind reversal above 10 hPa. However, it is unable to fully develop the waves (particularly wave number two) at 5 hPa, where the NMC data show the largest amplitude. This is also reflected in the thermal field, which should have briefly risen above 250 K at 3 hPa. Following this event, mean zonal winds recover only to 45 m sec$^{-1}$ at this latitude. While stratospheric temperatures up to this point have been well simulated at 60° N, they begin rising during March, slightly ahead of the observations. Lower stratospheric (30 hPa) temperatures are up to 5 K too warm in December and January. After the February warming they are cool, especially close to the reference surface. The declining wave amplitudes and the rising stratosphere temperatures indicate that the seasonal transition is beginning.

### Residual Mean Meridional Winds

Several studies have established the usefulness of the residual mean winds as a diagnostic for understanding transport in the meridional plane. Holton [1981] shows how the residual circulation from a three-dimensional model is comprised of the diabatic circulation plus effects from wave transience and dissipation. Mahlman et al. [1986] and Holton [1986] expand this concept by including discussions of the relative roles played by the diabatic circulation and quasi-horizontal mixing in determining the meridional slopes of the tracer isolines. In their GCM, Plumb and Mahlman [1987] find that the transport circulation is reasonably represented by the residual mean circulation. Recently, Douglass et al. [1993] provided a direct comparison of the zonal mean structure from a three-dimensional simulation with that from a consistently formulated two-dimensional model. Thus we will use the residual mean circulation from the GSMM to aid interpretation of its transport characteristics. However, we will also illustrate where the three-dimensional aspects of the circulation impose limitations to the application of this diagnostic tool.

The time-averaged components of the residual mean circulation for December and March are illustrated in the top and bottom, respectively, of Figure 7. These approximately represent solstitial and equinocial conditions of a Brewer-Dobson type circulation. In December the vertical component is positive over the tropics below 55 km and over most of the southern hemisphere. It is negative over most of the northern hemisphere, except at the pole, where the planetary waves have displaced the vortex away from the pole in the time average. With the meridional component positive nearly everywhere above 20 km, trace constituents initially lying over the south pole should be lofted, advected northward most vigorously near the model lid, and forced to settle over the north polar cap. That is, with the meridional speeds approaching 4 m sec$^{-1}$ near the model lid in the tropics, parcels originally positioned near the south pole and 0.10 hPa will appear in the northern hemisphere mesosphere after about five weeks.

In March, as the seasonal transition is under way, the vertical component becomes positive throughout the tropical model stratosphere and mesosphere. It remains negative in the time-averaged northern polar vortex and has switched sign over the south pole, where the southern polar vortex is strengthening. The meridional component is now negative south of 15° S. It is during February that the residual mean meridional circulation begins its seasonal transition, first in the southern hemisphere. The tracers respond to the seasonal transitions and show at which heights it first begins.

### Passive Tracer

A two-dimensional summary of the evolution of the passive tracer is given in Figure 8. The first panel illustrates the zonally averaged tracer mixing ratio fifteen days following initialization. The other panels show the configuration of this field at the beginning of each subsequent calendar month. Throughout the winter months the northern column descends while the southern column is bifurcated. In the southern hemisphere, material below 10 hPa is pushed downward and apparently some is lost below the reference surface. Above 1 hPa, a large portion of the material is advected into the opposite hemisphere after being pushed upward. Thus material which originates in the southern column appears north of the equator well before the end of the first month of the integration. After two months, some of this material is entrained into the northern polar vortex in the model's upper mesosphere and is substantially confined to the vortex as it descends. This pole-to-pole circulation is remarkably consistent with the stream function of the residual mean meridional circulation.

The structure in the zonal means near the south pole between 2 and 20 hPa reflects the transition of the circulation from its winter to summer regime. At pressures less than 0.4 hPa, the southern summertime anticyclonic vortex is established. However, in the upper and middle stratosphere, diabatic forcing is strengthening the anticyclonic circulation and displacing the weakening cyclonic polar vortex to more northerly latitudes. Accompanying this growth of planetary wave amplitude is cross-polar flow. The tracer is thus stretched and twisted. Occasionally a "clump" of material is rapidly advected over the pole. When this occurs, high values are only briefly seen in the zonal average since the tracer is then immediately advected equatorward. There is essentially no diffusive mixing responsible for the transience in the zonal means during these episodic events. Further, the planetary waves thrust air which has little or no tracer content from the middle and low latitudes toward the pole. This further promotes the meridional redistribution of the tracer.

A discussion of mixing during modeled final warmings is provided by Hess [1991]. Individual vortices are being sheared in the vertical. As they lose their identity, the tracer is then subject to rapid horizontal mixing. The two-dimensional cascade is faster for the tracer, so it becomes decorrelated from the increasingly shallow potential vorticity anomalies. This behavior is different from that seen near the north pole at this time,
Figure 7. The residual mean meridional wind from the mechanistic model, time averaged for (top) December and (bottom) March. (left) Meridional component (m sec$^{-1}$) and (right) vertical component (cm sec$^{-1}$).
Figure 8. Latitude-pressure cross sections of the zonally averaged passive tracer concentration on the indicated dates. Units are arbitrary and the contours are 0.5, 1, 2, 3, ..., 10.
since the northern polar vortex is forming. After a period of initial adjustment, the descending tracer from the northern column becomes aligned with the potential vorticity. Until the vortex is disturbed by wave events, the zonal means are steady and there is little mixing of middle latitude air with polar air.

The time tendency of the zonally averaged passive tracer at 80° S and 80° N latitudes is presented in Figure 9. Near the south pole, the bifurcation of the tracer at 10 hPa is evident after the spin-up period. Aloft, concentrations remain high until the mass traveling northward is greater than that being pushed upward. By January 1, 1990, the flow in the entire southern atmosphere above 10 hPa is easterly, circumpolar, and steady. Thus the transition from mean upward to mean downward vertical velocity can be deduced by the inference in the slopes of the mixing ratio contours. This begins in early February at 4 hPa and occurs later as a function of decreasing pressure. At 0.02 hPa it occurs around March 10, 1990.

During the southern atmosphere's transition from easterlies to westerlies, the rotation of the tracer in the southern upper atmosphere is dramatically upset. Until that time, it is in constant westward motion, concentric with the pole, and horizontally sheared, with the latitude of strongest easterlies well modeled at 15°-30° S, depending on pressure. At 0.10 hPa westerlies first appear at 60° S on February 26, 1990. Along this latitude circle, the tracer halts its westward motion and reverses direction. At both higher and lower latitudes the westward drift continues. One zero wind line rapidly propagates toward the pole and the other gradually propagates toward the equator. The tracer is strongly sheared at the zero wind lines. The potential vorticity develops a relative minimum which encircles the pole, so the latitudinal gradient is small. The tracer thus rapidly aligns itself with the potential vorticity during the transition. When the zero wind line reaches the pole, the vertical velocity at the pole becomes noticeably negative. At levels above ~10 hPa, higher tracer amounts begin to be transported downward. Animation thus shows the orderly change of spin followed by rising tracer concentrations at the center of the new and developing winter polar vortex.

Over the north pole, decreasing tracer concentrations are initially a result of mixing associated with the initial adjustment. The positioning of the vortex, and its greater zonal asymmetry at lower levels, leads to more dramatic sloshing below 1 hPa. Thus in the lower stratosphere the potential vorticity and the tracer become well correlated after about three weeks, as described by Mahlman and Morim [1978]. But in the upper stratosphere and mesosphere, correlations poleward of 60° N are greater than 0.90 after only ten days. Once the tracer is aligned with the potential vorticity and is essentially confined to or follows the movements of the vortex, animation indicates that descent becomes the primary cause of the decreasing mixing ratios. Remnants of tracer from the northern column descend below 0.10 hPa by November 20, 1989, and below 1 hPa by December 5, 1989 (Figure 9b). The rate of descent lessens through the winter.

The northern polar mesosphere is "clean" until material which originated in the southern hemisphere propagates into the region. It reaches 80° N near the model lid on December 6, 1989. Since the mean meridional transport is stronger near the lid, it takes the tracer longer to cross the equator at higher pressures. As the material attains more northerly latitudes, it begins drifting downward, reflecting the influence of polar night. At 0.10 hPa, material crosses the equator in the middle of December and is moving northward at approximately a degree of latitude per day. Animation shows that as the material appears on this pressure surface at the pole in late December, it is descending, and mixing ratios begin increasing rapidly since descent there is strongest. At middle northern latitudes, mixing ratios slowly increase as material aloft is slowly pushed downward. This behavior extends down to approximately 1 hPa (see Figure 8).

Care must be exercised before concluding that a bend of the zonally averaged tracer isolines indicates a directional change in the vertical component of the transport circulation. For example, the decrease in tracer content at 80° S between 20 and 70 hPa in mid-December (Figure 9a) is caused by the quasi-horizontal movement of tracer to more northerly latitudes. During this period the amplitude of geopotential wave-one does not exceed 300 m. In this case, vertical advection is not playing a major role. GSMM vertical velocities collocated with the tracer mass are generally of the order of 0.25 cm sec⁻¹, with occasional values briefly as high as 0.50 cm sec⁻¹.

In the northern hemisphere, large, periodic changes in the high-latitude zonally averaged tracer concentration (Figure 9b) indicate amplification and decay of the planetary waves. Rapid rises and falls are due to the latitudinal displacement of the vortex. As the vortex moves southward, the relatively large tracer which has been descending in its core is taken with it. Air from middle latitudes which contains smaller amounts of tracer takes its place. In reverse, the high-latitude contours rapidly rise when wave amplitude declines and the vortex moves back onto the pole. In February, for example, tracer amounts decline in the deep layer bounded by 70 and 0.02 hPa at 80° N.

In addition to the bodily movement of the vortex, both the horizontal and vertical advection during periods of high planetary wave amplitude must be considered. Strong winds on the equatorial side of the displaced vortex may penetrate the subtropics leading to rapid poleward transport of middle- and low-latitude air [Leovy et al., 1985, Treptol et al., 1993]. With the configuration of the passive tracer, cleaner air is thrust poleward and acts to further reduce the mixing ratios. On February 7, 1990, the vertical wind at 0.4 hPa approaches 12 cm sec⁻¹ at 75° N and 15° W, forcing air containing smaller concentrations of tracer to ascend. In fact, between February 4 and 12, 1990, the modeled 3000 K potential temperature surface rises from
Figure 9. Time-pressure diagrams of the zonally averaged passive tracer concentration for (a) 80° S and (b) 80° N latitude. Units are arbitrary and the contours are 0.5, 1, 2, 3, ..., 10.
47.5 to 51 km at this point. With the vortex displaced from the pole, the region of descent lies farther to the south. This spatial dependence of the vertical circulation demonstrates a limitation of using zonally averaged diagnostics to interpret the transport characteristics of large planetary wave events.

Nitrous Oxide

Figure 10 illustrates the progression of the zonally averaged $\text{N}_2\text{O}$. Unlike the passive tracer, the impact of the chemistry must also be considered (see Figure 4). But because of $\text{N}_2\text{O}$'s long lifetime in most of the stratosphere, several of the features which are seen in the evolution of passive tracer are replicated here. Most notable is the remarkable descent of the contours over the north pole during the first two months of the integration. At this time, the region is not directly influenced by the $\text{N}_2\text{O}$ destruction, and the downward drift of the contours is due to cooling in polar night. The rates of descent are largest in the mesosphere and upper stratosphere and are less in the lower stratosphere. But the slow descent of the contours in the lowest few levels continues throughout the integration, with the 260 ppbv contour falling below the reference surface poleward of 70° N in February.

In the mesosphere, the zonal averages show that the mean transport circulation has a significant influence even in the presence of the photochemical destruction. Near the model lid, where residual mean meridional wind speeds are greatest, the mixing ratios at most latitudes slowly rise as the contours bend northward until sometime in January. At 0.02 hPa the 0.20 ppbv contour migrates from 35° S at initialization to 50° N on January 1, 1990. Over the south pole, throughout November and early December, the contours are drifting upward above 10 hPa. But destruction increases as a function of height and is strongest near solstice. So poleward of 60° S and at pressures less than 0.3 hPa, where $\text{N}_2\text{O}$ densities are already small, mixing ratios are less on January 1, 1990, than at the start of the previous month.

As the seasonal transition begins in February, mesospheric $\text{N}_2\text{O}$ which lies north of the equator continues heading toward the north pole, while the material (of higher mixing ratio) which remains in the southern hemisphere begins to move southward. This meridional flux divergence, along with larger photochemical loss rates in the southern mesosphere which act to reduce the higher concentrations there, creates a homogeneous $\text{N}_2\text{O}$ layer in the mesosphere by March. During March the photochemistry continues to slowly decrease the mixing ratios in this layer. Over the south pole the 0.50 ppbv contour drops from 0.01 to 0.40 hPa. Strengthening mean polar descent in the mesosphere and upper stratosphere is an important component to the downward drift of the contours as equinox approaches and as photochemical destruction slows over the south pole.

Over the tropics the $\text{N}_2\text{O}$ contours are moving upward between 20 and 1 hPa from initialization until the end of January. The zonally averaged mixing ratio at 1 hPa increases from 5 to ~20 ppbv at the equator. The steady ascent creates large meridional gradients in the CTM's low latitudes in this layer.

A clearer picture of the evolution of the polar zonal means of $\text{N}_2\text{O}$ is provided by the time series illustrated in Figure 11. At 80° S at pressures less than 1 hPa, the dynamical forcing appears to dominate the photochemistry until early December. That is, mixing ratios are rising because of relatively strong transport from below. By solstice, however, the photochemistry is dominant between ~0.3 hPa and the lid, and the mixing ratio falls below 2 ppbv. At pressures greater than 1 hPa, the change in mixing ratio is small until the middle of February, indicating a balance between the chemical and dynamical tendencies. By February, and especially during March, the loss rates are declining and the dynamical forcing begins to dominate. By equinox, in spite of nearly zero loss rates, the mixing ratios are falling. This is in response to the poleward and downward transport of depleted $\text{N}_2\text{O}$ in the developing southern polar vortex.

Near the north pole the evolution is much more irregular due to the influence of amplifying and decaying planetary waves (Figure 11b, see also Figure 6). Until December the wave events are relatively small, and the $\text{N}_2\text{O}$ contours steadily descend. But afterward, the perturbations are large enough to push the vortex off the pole. As the vortex migrates southward, the depressed mixing ratios are carried with it and the high-latitude zonal averages rise as middle latitude air takes its place. Additionally, the large wave events tend to capture tropical $\text{N}_2\text{O}$ in the region of strong winds on the equatorward side of the displaced vortex. This further enhances the high-latitude zonal means if the tropical air is thrust near the pole. With midwinter events, strong radiative forcing encourages the decay of the temperature perturbation. As the wave amplitude declines, the vortex boundary migrates back toward the pole, bringing the lower mixing ratios with it.

Thus $\text{N}_2\text{O}$ at high latitudes may rapidly rise and fall. This is particularly evident in early and late December, respectively, at 1 hPa. Also, between February 1 and 5, 1990, at 2 hPa, the mean ratio rises from 1 to 10 ppbv. However, a major difference between the two events is that the former occurs while the vortex is at midwinter strength and retains a distinct boundary. When the vortex moves back onto the pole, the mean $\text{N}_2\text{O}$ contours return nearly to their pre-event pressure levels. The transport seems nearly reversible since a rather symmetric vortex can again become established. The February event, on the other hand, occurs near the end of winter, prompts a wind reversal, and may be considered the final warming. A winter-like vortex does not again form and mean zonal winds do not again rise above 45 m sec⁻¹. The vortex wobbles back onto the pole only briefly, and afterward, especially in the upper stratosphere and lower mesosphere, the mean contours are at pressures considerably less than they were before the event. In this case, the transport is irreversible.
Figure 10. Latitude-pressure cross sections of the zonally averaged N₂O mixing ratio on the indicated dates. Units are ppbv and the contour interval is nonuniform.
Figure 11. Time-pressure diagrams of the zonally averaged N$_2$O mixing ratio for (a) 80° S and (b) 80° N latitude. Units are ppbv and the contour interval is nonuniform.
Figure 12 illustrates the evolution of N$_2$O on the 1000 K potential temperature surface (32-34 km) during the middle of February. On February 11, 1990, tropical N$_2$O is rapidly transported over northern Alaska after being captured by the winds on the equatorward side of the displaced vortex. Four days later, with most of the tropical source cut off, the high mixing ratios appear in two distinct centers, with one over Scandinavia and the other over Canada. The former continues rotating around the cyclonic vortex and loses its identity after another four days. The larger mass drifts southwestward under the influence of the anticyclone and is located near 40° N and 140° W on February 19, 1990.

Nearer the pole, mesospheric values of N$_2$O, less than 1 ppbv, are visible on February 11 and 15, 1990, at 70° N and 135° W. This reflects the position of the polar vortex that has been displaced from the pole during the warming. On February 11, 1990, a long tongue of depleted N$_2$O, stripped from the polar vortex, stretches across the Atlantic at about 30° N. A large parcel of low N$_2$O air has been disconnected from this tongue and resides over the Pacific, drifting westward on the equatorward side of the anticyclone. The gradients between this southwardly displaced vortical air and the subtropical N$_2$O are large. Figure 12 thus shows the ability of the modeled planetary waves to rapidly transport air from the subtropics to polar latitudes and, at the same time, maintain a clearly identifiable pocket of polar air. While there is significant mixing in the surf zone, N$_2$O concentrations in the middle latitudes are by no means horizontally homogeneous. The planetary-scale anticyclone, in this case, retains its isolation with some robustness.

Conclusion

In this paper we have presented the results from experiments using a GSMM. We forced the lower boundary of the model with observed 100-hPa heights and placed the lid at 0.01 hPa. Tropospheric processes are not explicitly included. The GSMM, which has an accurate radiative transfer scheme, was initialized on October 20, 1989, and was integrated through April 1, 1990. In the southern hemisphere, this time interval included the breakup of the westerly polar vortex during November, the transition to summer easterlies, and the formative stages of the next westerly vortex in March. In the northern hemisphere the dynamics were dominated by the wintertime polar vortex and the continuous influence of planetary waves from December through February. During February a final warming occurred. However, the transition to the summertime regime was not complete at the end of the integration.

![Figure 12. North polar projections of the N$_2$O mixing ratio (ppbv) on the 1000 K potential temperature surface on the indicated dates. The contour interval is nonuniform and mixing ratios above 50 ppbv are not shown.](image-url)
Winds and temperatures from the GSMM were compared with NMC observations where possible. The stratopause temperature in both hemispheres became biased shortly following the initialization, with the summer pole becoming too warm and the winter pole becoming too cold. Throughout the first month, there was a high degree of consistency between the observed and modeled variability. After the first month, however, the day-to-day similarity between the model and the observations decreased. Nevertheless, the model simulated most wave events in the northern winter. Therefore, the long-term variability qualitatively followed that in the observations. The influence of the radiative heating, along with the lower boundary forcing, ensured that the seasonal transitions occurred at the correct times.

A primary objective for developing the GSMM is to use it as a tool to study stratospheric and mesospheric transport. The winds and temperatures from this five-month experiment were interfaced with an offline CTM, and two transport experiments were performed. In the first, a conservative tracer, which approximated the perturbed NO\textsubscript{2} distribution (but not its chemistry) following an SPE, was initialized in columns over the geomagnetic poles on November 1, 1989. This experiment demonstrated how the GSMM's winds transport mass in the meridional plane. Material which was placed over the north pole rapidly descended because of the influence of the strong radiative cooling. Air which was originally located in the northern polar vortex at 65 km descended to 30 km by the middle of January and was replaced by clean air which was originally in the mesosphere at middle and low latitudes. Over the south pole, tracer material at pressures less than 10 hPa was advected upward and northward in the mesosphere. Within 30 days some of this southern hemisphere material could be found at the north pole, and within 60 days some was descending into the northern polar stratosphere. This suggests a substantial interhemispheric mass flux, on a seasonal scale, at pressures less than 1 hPa. As the high-altitude air converges into the wintertime polar vortices, it rapidly descends into the stratosphere. Fisher and O'Neill [1993] find strikingly similar meridional transport in their study of parcel trajectories.

Jackman et al. [1993], when using these GSMM winds and a similarly configured NO\textsubscript{2} perturbation, have recreated several aspects of the observed ozone depletions following the October 1989 SPEs. In particular, the hemispheric asymmetry of the depletion and its longer persistence over the Arctic than over the Antarctic was simulated. Examination of the transport from that SPE experiment, along with the results presented here, suggests that the evolution of the perturbation was strongly related to the dynamics at the time of the event. In the southern hemisphere, the polar vortex was breaking up when the SPE occurred. With the breakup, unperturbed (low NO\textsubscript{2}) air from the middle and low southern latitudes was irreversibly mixed with the perturbed (enhanced NO\textsubscript{2}) high-latitude air. There, ozone reduction resulting from the SPE was relatively short-lived. In the northern hemisphere the NO\textsubscript{2} perturbation was eventually confined to the developing polar vortex. As the vortex moved into and out of sunlight, ozone was destroyed and the interior of the vortex was gradually filled with the ozone-depleted air. By inference, it is likely that larger localized ozone depletions would have been possible in the southern hemisphere if the SPE had occurred earlier and had been confined to the still mature polar vortex. If the SPE had occurred later, after the establishment of the summer easterlies, latitudinal redistribution of the perturbed constituents would have been much slower because of the lack of planetary wave activity in that regime. We would also expect a general tendency for the perturbation to drift upward into the mesosphere. Hence a summertime SPE would generate a substantially different ozone depletion signal which could, quite possibly, be accurately simulated with a two-dimensional model.

In the second transport experiment a realistic and balanced distribution of NO\textsubscript{2} was initialized on November 1, 1989, and a parameterized photochemical destruction was included. Since the time scales for the destruction are fastest in the mesosphere, it was more difficult to demonstrate the scale of the interhemispheric transport that was seen from the passive tracer. Over the poles, however, the NO\textsubscript{2} contours showed significant descent during winter at all levels and ascent during summer in the upper stratosphere and mesosphere. Over the tropics, NO\textsubscript{2} mixing ratios were generally increasing as a result of upward advection by the mean meridional circulation. The wave events which prompted the migration of the vortex away from and toward the pole could be traced in rapid rises and falls, respectively, of the zonally averaged NO\textsubscript{2} contours at high northern latitudes. Even though substantial descent was modeled in the winter hemisphere, the slopes of the zonally averaged contours in the lower stratosphere implied that the simulated rates of descent are less than in the aircraft observations from AAOE and AASE. The mechanistic approach thus provides improved representation of wave activity in the stratosphere compared to low-resolution GCMs. The “cold” pole problem and the associated underestimation of descent rates appears to be less severe but is not eliminated.

Analyses of observations and of this and other recent three-dimensional modeling results [Fisher and O'Neill, 1993, Mote et al., 1993] are contributing to an emerging picture of descent within the winter polar vortex. Russell et al. [1993] and Tuck et al. [1993] show that the UARS Halogen Occultation Experiment (HALOE) detected mesospheric values of methane (CH\textsubscript{4}) at pressures as high as 40 hPa within the Antarctic vortex, indicating, in their words, deep, unmixed descent. NO\textsubscript{2} should behave similarly. From our simulation of the February warming, Figure 13a indeed shows the distinct infusion of mesospheric air down the core of the displaced polar vortex. In fact, the contours indicate descent down to the reference surface. Two days later, Figure 13b shows the zonally averaged 5-ppbv NO\textsubscript{2} contour cut off at 10 hPa. Horizontal projections indicate that the core of the depleted NO\textsubscript{2} has moved back onto the pole in the lower and middle stratosphere. In the
upper stratosphere, meanwhile, the wave amplitude is still large enough to keep the vortex away from the pole. This should be a typical signature during winters which are wave-one dominated and is one which cannot be explained with purely zonally averaged transport.

The model simulates the unmixed descent observed by HALOE but within a vortex that is often displaced from the pole. Descent begins before equinox during the vortex’s formative stages and is steady throughout autumn while the wave amplitudes are weak. The picture then becomes complicated during winter by planetary waves which transport tropical air to high latitudes. Leovy et al. [1985] confirmed the extent of such meridional transport in LIMS ozone observations. The warm air which remains at polar latitudes must then descend as it radiatively cools. However, our simulations indicate that some of the low- and middle-latitude material ends up circulating around the anticyclone in strong wave-one events. For N$_2$O, this is revealed in time series as an upward bulge in the high-latitude zonal mean contours.

An important question, which we expect to investigate by lengthening these experiments, is the degree to which mesospheric air is “trapped” in the lower stratosphere after the final warming. In November and December over the south pole, as the summertime easterlies became established, the conservative tracer at pressures less than 10 hPa ascended into the mesosphere, while the tracer which was at greater pressures continued to descend. If, during the winter, this is preceded by deep, unmixed descent, then some air of mesospheric origin should remain at low levels. Air which has descended only to the upper and middle stratosphere appears to simply ascend back into mesosphere and becomes available for the solstitial interhemispheric transport. The details and timing of the final warmings may determine how much air can be processed between the mesosphere and the stratosphere.

Certainly, the mean and residual circulations are useful and important tools which aid diagnosis of the atmosphere’s transport characteristics. However, these experiments demonstrate that care must be taken when using zonally averaged illustrations. Even weak planetary waves are able to induce transport in the horizontal plane. If the longitudinal averaging is done close to the pole, then the contours may move vertically even though only negligible vertical motions are involved. Stronger planetary waves will often transport large quantities of equatorial material poleward. If the meridional gradient of a constituent is sufficient, and if some of the equatorial air passes close to the pole, then the zonally averaged constituent contours will likely exhibit transient variability. Finally, if the pole vortex is tilted with height and sits on the pole in the lower stratosphere, then long-lived trace constituents which have descended in its core will appear cut off in zonally averaged latitude-height cross section plots. To explain this signature, reference only to the mean circulation is insufficient.

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